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Late Quaternary Meltwater Pulses and sea level change

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ABSTRACT

After the Last Glacial Maximum (LGM) global mean sea level (GMSL) rise was characterised by rapid increases over short (decadal to centennial) timescales superimposed on a longer term secular rise. These rapid departures from the long-term trend have been termed Meltwater Pulses (MWP) and understanding the nature and timing of these provides insights into past periods of GMSL response when climate forcing was high and, by analogy, the possible nature of future responses. However, at present great uncertainties exist in our knowledge of MWPs, where they originated and the relative contribution of drainage of ice-proximal and pluvial lakes and discharge into the oceans from disintegrating ice sheets. This paper is a contribution to our understanding of the extent and significance of MWPs.

see some changes in wording

Key Words: Meltwater Pulse, Last Glacial Maximum, Global Mean Sea Level

INTRODUCTION

In past periods of deglaciation sea level rise has been characterised by rapid increases over short (decadal and centennial) timescales superimposed on a longer term secular rise. During the late Quaternary these rapid departures from the long-term trend have been termed Meltwater Pulses (MWP) and understanding the nature and timing of these provides insights into past periods of sea level response when climate forcing was high (Dutton et al 2015) and, by analogy, the possible nature of future responses. However, at present great uncertainties lie in our understanding of MWPs, where they originated and the relative contribution of drainage of proglacial and subglacial lakes and discharge of ice into oceans from disintegrating ice sheets. In addition, some postglacial rapid sea level rise events have not been designated widely as MWPs despite their being strong candidates. In this paper we assess these uncertainties, review the literature on MWPs and discusses their origins, magnitude, extent and impact on coastal and submarine processes. We produce the first quantification of the Global Mean Sea Level (GMSLR) rise associated with lake drainage and show that this lake drainage resulted in about 1% of the total GMSLR from the end of the LGM to the mid-Holocene. We argue for an early Holocene MWP, and end by assessing current attempts to improve our understanding of MWPs. All dates are expressed in calendar years before present, thus 19.5ka BP. In tables 1 and 2, dating sequences involving U/Th are indicated.

DECAY OF GLOBAL ICE SHEETS AFTER THE LGM

During the LGM, the major global ice sheets extended well beyond present coastlines, and their impact on both GMSL and glacial isostatic adjustment (GIA) is well known. In some areas, where glacio-isostatic depression of the land beneath and surrounding the ice sheets did not exceed GMSL fall, the ice ended on land; in others, where GMSL fall was insufficient to reveal the surrounding sea floor, ice sheets ended in floating ice shelves, although these effects were locally modulated by the effect of the reduction in the gravitational attraction of the ice mass as it retreated. As decay progressed and GMSL rose, ice shelves increased in extent as the rate of GMSL rise exceeded that of glacio-isostatic uplift. Recent summaries (e.g. Dyke et al., 2002; Svendsen et al., 2004 Jakobsson et al., 2014; Stroeven et al.,2016; Patton et al., 2016) outline particularly extensive ice shelves along the margins of the Barents-Kara, Fennoscandian and British-Irish (Celtic) ice sheets; the Laurentide and Innutian ice sheets; the Cordilleran Ice sheet and both East and West Antarctic ice sheets. In contrast, the Siberian and Patagonian ice sheets only locally produced ice shelves. The break-up of ice shelves probably produced the greatest volume of water fuelling the rise in GMSL, but other sources were the discharge of meltwaters, impounded pro-glacial lakes and in areas beyond the ice, pluvial lakes.

It would be surprising if the varied topography beneath the decaying ice sheets combined with variations in the topography of the ocean floor and changes in both global and regional climate should not have resulted in variations in the pattern and rate of deglaciation and consequent variations in the supply of meltwaters to the oceans. This consideration has been a driver in attempts to identify meltwater pulses.

THE RISE IN GMSL AFTER THE LGM

The LGM is dated globally at between 26,000 and 19,000 calendar years before present (cal. BP) (e.g. Clark et al., 2009), and the rise in GMSL thereafter has been widely documented and modelled (e.g. Lambeck et al., 2014; Peltier et al., 2015). In general terms, the rise was at first relatively rapid, averaging 10mm/year, but by c. 7000 cal. BP it had slowed markedly to less than 1mm/year (e.g. Nakada and Lambeck, 1988). The rate may then have fallen to below 1mm/year (Kopp et al., 2016) until the recent acceleration to around 3mm/year associated with anthropogenic global warming (AGW) (Figure 1). MWPs occurred against this background of GMSL change.

MWPs are commonly viewed as global phenomena involving the rapid introduction of meltwater into the ocean from ice-proximal environments as ice sheets decay and ice-dammed and subglacial lakes drain, as illustrated by Carlson and Clark (2012). Table 1 lists MWPs recognised in the literature according to designated name, age, origin and location affected, while Figure 2 shows where they have been identified. The sequence we have developed is based on published correlations (where no correlation is given the event is regarded as separate). It should be noted that several MWPs are contentious or uncertain, while others are only recognised regionally or locally. Thus MWPs 2A and 2B of Liu et al. (2004), based on evidence from the South China Sea and correlated with evidence elsewhere in the western Pacific are apparently not specifically recognised in subsequent literature; MWP1B is only recognised in the Caribbean, specifically Barbados (Fairbanks, 1989; Fairbanks et al., 2005, Peltier and Fairbanks, 2006, Abdul et al., 2016) and the western Pacific (Liu et al., 2004) and is believed uncertain in Tahiti (Bard et al., 2010, 2016, Mortlock et al., 2016); MWP1C (Liu et al., 2004) is not supported elsewhere, and the MWP associated with the 8.2 event is probably distinct from the MWPs dated at ~7.5ka. cal. BP although the two episodes are combined as MWP3 in Gornitz (2013). We suggest that the MWPs between c.11.5ka. cal. BP and c. 7ka. cal. BP are probably difficult to differentiate globally given that they occur at the time of rapid GMSL rise of 12-13mm/year (Smith et al., 2013) following the end of the Younger Dryas (c.11,700 cal. BP) and before c.7ka cal. BP: the early Holocene sea level rise of Smith et al. (2011).

The initial effects of the release of meltwater were probably registered rapidly across the oceans, although the full effects may have taken much longer, thus in modelling the effects of an instantaneous thinning of the West Antarctic Ice Sheet, Clark and Lingle (1977) found that far-field submergence would have been “immediate” (before slow emergence of these areas on a centennial scale took place as sea water was redistributed). Some of the MWPs are believed to have involved very rapid rates of GMSL rise (> 45mm/yr) yet only MWP 1Ao, MWP 1A, MWP1B and the MWP in the early Holocene after the final c.8470 cal. BP floods from the Lake Agassiz-Ojibway system, have been claimed to have been more than regional events.

MWPs AND ICE SHEET DISINTEGRATION

The causes of MWPs have been widely discussed, as the debate on the origins of MWP1A (e.g. Clark et al., 1996; 2010; Peltier, 2004, 2005; Carlson and Clark, 2012) shows.

Sources of meltwater from ice sheets. The major source of MWPs is probably to be found in episodes of rapid disintegration of the Late Quaternary ice sheets as global climates ameliorated (MacAyeal 1993). The timing of such episodes would not have been synchronous, but would have depended upon regional climate, local subaerial and subglacial topographic changes in relative sea levels (reflecting the interaction of GMSL and GIA) as well as regional sea water temperature. The greatest sources of meltwater would have been from the decay of those ice sheets terminating on continental shelves. However, the rate at which these meltwaters discharged into the oceans is unclear.

North America. In North America, Hendy and Cosma (2008) commented upon the rapid retreat of the Cordilleran Ice Sheet from the LGM as demonstrated by ice rafted debris offshore, so that by ca. 10.8ka. BP ice was no more extensive than today (Clague and James, 2002) and meltwater delivery uncertain. To the East, the Innutian ice sheet (Dyke et al., 2002), confluent with the Laurentide and Greenland ice sheets and occupying much of the Canadian Arctic archipelago, may have remained at its maximum extent until after the LGM before ice sheet thinning and ice margin recession occurred widely after circa 16ka. cal. BP, except to the North, where recession occurred much later, and thus may not have contributed to late Quaternary meltwater pulses before MWP1A. Contributions to later MWPs may have been provided from rapid retreat of remains of the Innutian Ice Sheet for example on southern Ellesmere Island during the Holocene at around 9.3ka. cal. BP (England et al., 2004). From the Laurentide Ice Sheet~~,~~ the proposal of Shaw (1989), Blanchon and Shaw (1995) and Rains et al. (1993) that subglacial meltwaters in the area of Alberta drained southwards into the Mississippi is contentious. They argued that meltwater originating in the Livingstone Lake area, Alberta, drained sometime between 18ka. and 15ka. cal. BP, with a volume of 48,000 km^3 and amounting to 0.23m of rise in GMSL. Clarke et al. (2005); Benn and Evans (2005); Evans et al. (2006) have strongly doubted whether this occurred at all. However, to the East, meltwaters from the Hudson's Bay area and St. Lawrence valley probably flooded rapidly into the North Atlantic. Here, Marshall and Koutnik (2006) summarised rapid freshwater input into the North Atlantic according to several authors. They identified GMSL rise in Heinrich events ranging from as little as 0.39m to 3.9m (Dowdeswell et al., 1995) to 10-20m (Hemming, 2004), depending upon the length of the event. In Greenland, the initial ice sheet retreat from the LGM limit began as early as 17,000 cal. BP in some areas occurred after early MWPs. Funder et al. (2011) argue that break-up of this ice sheet on the continental shelf was due to several factors, such as sea level change, ocean warming, shelf bathymetry and that initial recession was by calving.

Eurasia. In Eurasia, the break-up of the smaller British-Irish (Celtic) Ice Sheet, surrounded by an extensive continental shelf, was particularly marked around 17ka. – 16ka. cal. BP (Clark et al., 2012), coincident with Heinrich 1 but before the Killard Point Stadial (McCabe and Dunlop, 2006), Wester Ross Readvance (Ballantyne et al., 2009), and probably the Perth Readvance (McCabe et al., 2007). The subsequent decay of ice after these events may have contributed to MWP 1A. From the southern margin of the Fennoscandian Ice Sheet, proglacial Lake Disna developed after the LGM and repeatedly flooded into the Denipr catchment in a series of pulses into the Black Sea between 20ka. cal. BP and 16ka. cal. BP and ultimately into the NE Mediterranean (Soulet et al., 2013), thus possibly contributing to an early MWP. Further North, Rinterknecht et al. (2006) showed that rapid deglaciation occurred at around 16ka. cal. BP when the ice sheet in the SE retreated along the present Polish and Lithuanian coasts before later readvancing after circa 15,500 cal. BP and then retreating again circa 14.6ka. cal. BP, an age coincident with MWP 1A. Nesje et al. (2004) remark that freshwater outbursts from the Baltic Ice Lake occurred at the same time as outbursts from the Laurentide Ice Sheet, thus increasing freshwater input to the North Atlantic. In the Arctic,~~,~~ the Barents-Kara ice sheet, confluent with the Scandinavian ice sheet, covered large areas of the extensive continental shelf off western Siberia at the LGM, extending northward to Svalbard and eastward to Novaya Zemlya, and was up to 2000m (Landvik et al., 1998) or even 3000 m (Lambeck, 1996) thick. The shelf topography beneath the Barents ice sheet was influential in its decay. The northern margin had begun to retreat from the shelf edge by circa 17.5ka. cal. BP at Svalbard (Landvik et al., 1998) and offshore Franz Josef Land at circa 18.7ka. cal. BP (Kleiber et al., 2000) about the time of Heinrich 1. Lucchi et al. (2015) identify the rapid deposition of interlaminated plumites (turbid meltwater deposits) in Storfjorden and Kevithola, Spitsbergen, which they date at 14.65ka.-14.31ka. cal. BP and interpret these as a sedimentary signature of MWP 1A. Otherwise few observations have been recorded on retreat stages which may have been associated with MWPs.

Antarctica. From Antarctica, rapid releases of meltwater may have contributed to major rises in sea level during MWP-1A and during MWP-1B. During the largest and most rapid of these, MWP1A, rates of GMSL rise reached approximately 4 m per century (Hanebuth et al., 2000; Peltier and

Fairbanks, 2006; Deschamps et al., 2012). Interpretation and modelling of far-field sea level records suggests a significant or dominant Antarctic contribution (Clark et al., 2002 and Weaver et al., 2003). Rates of meltwater contribution from the Antarctic Ice Sheet and the timing of former meltwater events are derived from marine geological records. Weber et al (2014) presented a detailed record of iceberg-rafted debris from the Scotia Sea reconstructed using coarse-grained sediments deposited from icebergs. The highest iceberg-rafted debris flux occurred c. 14.6ka. cal. BP, providing the first direct evidence for an Antarctic contribution to MWP1A. For Antarctica to have made a meaningful contribution to deglacial meltwater pulses MWP1A and MWP1B, Golledge et al. (2014) note that three conditions must be satisfied. First, sufficient GMSL equivalent ice volume must have existed in Antarctica at this time. Second, this ice volume must have been discharged at the correct time and at a rate fast enough to contribute to the observed rapid rises in sea level. Third, a plausible mechanism must exist to satisfy the previous two conditions.

Estimates of the contribution of Antarctica to MWP1A and MWP1B come from estimates of former ice-sheet thickness (and therefore assumed to be former volume) based on cosmogenic nuclide dating and numerical models of former ice sheet volume. Unfortunately, surface expose dating of glacial erratics, nunataks and other formerly glaciated surfaces using cosmogenic nuclides provide conflicting evidence. The evidence suggests that in some places the ice sheet was much thicker at the LGM and in others it was less so (Bentley et al., 2014) For example, Bentley et al. (2010) demonstrated that ice sheet thickness in the Weddell Sea at the LGM was thinner than previously suggested, with progressive thinning of the ice sheet by 230–480 m since ca. 15.0ka cal. BP. Ice volumes added 1.4 to 2.0 m to postglacial sea-level rise and would not have been sufficient to contribute significantly to MWP1A. Overall interpretation of Antarctic glacial geology from around the continent suggests only a very minor contribution (Licht, 2004, Bentley et al., 2010; Mackintosh et al., 2011). In their recent review of Antarctic deglaciation history, Bentley et al. (2014) found no evidence for major thickness changes at the time of MWP1A. They concluded that even after taking dating uncertainties into account this is consistent with only a minor contribution of Antarctica to this meltwater pulse. There also appears to be a consensus that the amount of thinning of the East Antarctic Ice Sheet is too small at 50 to 200 meters and likely too gradual and too late to have contributed any significant amount meltwater to MWP1B.

Using a numerical ice sheet model, Golledge et al. (2014) calculated that the maximum ice volume of the LGM Antarctic Ice Sheet was 32.3×10^6 km^3 by 20 ka cal. BP, ~5.8×10^6 km^3 greater than that of the modern ice sheet. This represents a eustatic s.l.e. lowering of ~14.5 m. Their models show rapid mass loss between 16.5 and 8.5 ka cal. BP, with fastest rates occurring during two main events at

14.5–14 and 11.6–10.2 ka cal. BP, and a smaller precursor event at 16.2–15.2 ka cal. BP. The early phase of ice loss, coincident with MWP1A, mostly affected the Antarctic Peninsula and the mid to outer Weddell Sea, with a small amount of thinning and grounding-line retreat also taking place in the central part of the outer Ross Sea. The second, more prolonged period of discharge (synchronous with MWP1B) took place primarily in the inner parts of the major marine embayments of the Weddell and Ross seas, and along the length of the Amery Trough. Their modelled ice sheet had sufficient excess ice volume to contribute significantly to deglacial meltwater pulses, and the predicted episodes of fastest mass loss are consistent with the timings of MWP1A and MWP1B, contributing around 2 m GMSL equivalent over the likely 350-year timeframe of MWP1A. During each of the two broad periods spanning MWP1A and MWP1B, their model simulations suggest ice loss from Antarctica ranging from 1.7 to 4.3 m GMSL equivalent. Other model results, however, contradict this. The model of Bassett et al. (2007), based on a subset of a suite of seven viscoelastic earth models and six different Antarctic deglaciation histories, indicated that the GMSL data do not rule out a large Antarctic source for this event. Their analysis indicates that the Weddell Sea was the most likely source region for a large (~9 m) Antarctic contribution to MWP1A. The Ross Sea is also plausible as a significant contributor (~5 m) from a GMSL perspective, but field data are not compatible with such a large and rapid melt from this region (Bassett et al., 2007).

The Global picture. From a global perspective, Carlson and Winsor (2012) have drawn attention to the difference between marine-based and land-based ice sheets, suggesting that in the Northern Hemisphere the southern margins of land-based ice sheets responded rapidly to climate warming, but that northern margins and shelf-based ice sheets showed a delayed response after which these ice sheets then responded catastrophically. The available data on ice sheet recession do not support any MWP other than 1A. We argue that nevertheless the circumstances of the circum-Arctic ice sheets predisposed them to widespread and relatively rapid decay after the LGM. The topography of the shelf areas would have precipitated rapid break-up of those areas where the overlying ice sheets occupied locally deeper areas, thus accelerating retreat and the production of meltwater and icebergs. As the receding grounding ice margin encountered locally deeper areas of shelf the pattern of retreat changed and local disintegration of ice streams would have promoted the periodic increase in iceberg fluxes. Recent research (Gomez et al 2015; Gregoire et al 2016) used numerical modelling experiments to simulate the extent to which the Laurentide ice sheet could have contributed to Meltwater Pulse 1A. Ice sheet perturbed physics ensembles were run to account for model uncertainties, constraining ice extent and volume with reconstructions of 21ka. years ago to present. They show that the North American ice sheet produced 3–4 m global mean sea level rise in

340 years due to the abrupt Bølling warming, but this response is amplified to 5–6 m when it triggers the ice sheet saddle collapse. They argue that the North American ice sheet could have contributed to MWP1a through two different mechanisms: the ice saddle collapse caused by the separation of the Cordilleran and Laurentide ice sheets (described in Gregoire et al 2012) and accelerated melt from the abrupt Bølling warming in the Northern Hemisphere at 14.6 ka. The separation of the two ice sheets on its own can produce a meltwater pulse of 5.7–11.0 m GMSL in 340 years associated with the saddle collapse. Thus, these results strongly suggest that the Bølling warming triggered the saddle collapse mechanism contributing 5–6 m or more to MWP1A from the North American ice sheet.

A similar situation probably obtained in Antarctica, where the East Antarctic ice sheet reached a mid-shelf limit at the LGM, while the West Antarctic Ice Sheet, reaching to the shelf edge, encountered a more variable bathymetry with the potential for rapid ice loss caused by ice-shelf break-up and acceleration and thinning of ice streams as is happening today across similar topographies closer to the continental margin (e.g Scambos et al., 2004; Rignot et al.; Joughin et al.). There is consensus that the most likely contribution of the AIS to rapid sea-level rise in the future is the rapid collapse of the West Antarctic Ice Sheet. Much of the West Antarctic Ice Sheet is grounded below sea level, with a steep reverse bed slope at the grounding line (Jamieson et al., 2012; Ross et al., 2012). Continued rapid recession of West Antarctic outlet glaciers therefore has the potential to contribute up to 5 m of sea-level rise over the coming centuries.

The principal contribution to GMSL rise after the LGM was from ice sheet flow off the land enhanced by the loss of buttressing ice shelves as they broke up and disintegrated. The rates by which sea levels rose may have exceeded initial rates of glacio-isostatic uplift and further increased the break-up of shelves in a feedback effect, enhanced by rises in sea water temperature. Where retreating grounding lines lay across a flat shelf or coastal area or in particular where a reverse slope was encountered inland, retreat was enhanced. We note that no major MWPs are registered after circa 7.6ka. BP, by which time the Laurentide, Scandinavian and north Asian ice sheets had virtually disappeared. Global temperatures had ceased to increase rapidly and most of the factors which led to the generation of MWPs were no longer present.

THE CONTRIBUTION OF PRO-GLACIAL AND PLUVIAL LAKES

Whilst the major contribution to MWPs was that of ice sheet disintegration, Interest has centred upon the effect of the rapid discharge of ice-proximal and pluvial lakes. Table 2 lists all published rapid discharges of meltwater from ice-proximal and pluvial lakes since the LGM as presently known. Figure 2 shows the locations of these lakes and the likely directions of meltwater flow.

North American Lakes. The largest lake systems draining into the ocean (partially beneath ice and along river systems) were in western North America, notably along the Columbia River, where Lake Missoula (Figure 3), impounded by an arm of the Cordilleran ice sheet, periodically broke through the ice dam and flooded through the channelled scabland beyond between 23ka. and 13ka. cal. BP (Bretz, 1969; Clark et al., 1984; Wiatt, 1985; Atwater, 1984, 1987; O'Connor and Baker, 1992; Benito and O'Connor, 2003). Together with the discharge of pluvial Lake Bonneville to the south at c. 17.4ka. cal. BP (e.g. Jarrett and Malde, 1987) along the Snake River the floods along the Columbia River system involving a total of up to 250,000 km^3 flowed into the Pacific, where sediments believed to have been deposited during these floods have been identified over large areas of the ocean floor (e.g. Brunner et al., 1999; Lopes and Mix, 2009), at least 400km beyond the mouth of the Columbia River. To the east, the Lake Agassiz-Ojibway system, which involved up to 205,000 km^3 in total discharged repeatedly between 12.94ka. and 8.47ka. cal. BP; the two last discharges, beneath the Laurentide Ice Sheet into Hudson's Bay having produced by far the largest floods (Teller et al., 2002; Teller and Leverington, 2004). Evidence indicates that the circa 8470 BP discharges from Lake Agassiz-Ojibway (Figure 4) may have been registered globally (see Table 2).

***South American lakes*.** In southern South America large lakes coalesced to the east of the Late Quaternary Patagonian Ice Sheet and at their maximum extent these had a surface area of ~7400 km^2 and an estimated volume of ~1500 km^3 (Glasser et al. 2016). They drained several times between 13,000 and 8,000cal. BP and the volume of water released during these drainage events was between 160 km^3 and 1150 km^3. Across the Atlantic, in Scandinavia, the Baltic Ice Lake and Ancylus Lake (Figure 5) discharge values are unclear, but involved perhaps c. 20,000 km^3.

Total volumes. The list of lakes in Table 2 does not include the several lakes formed as meltwaters drained southward from ice sheets in northern Eurasia, or pluvial lakes in Australia, since these drained into inland basins. The total volume of water discharged into the oceans as published amounts to c. 470,000 km^3, of which by far the greatest element was from lakes in North America. However, in most cases, discharge involved numerous separate events, sometimes over many decades or even millennia, and hence the immediate GMSL rise caused would have been less than is implied by the total volumes estimated. Nevertheless, in the case of the drainage from the Lake Agassiz-Ojibway system, where Teller et al. (2002) and Teller and Leverington (2004) identified 11 drainage episodes, modelling by Kendall et al. (2008) estimates that the last two floods accounted for 0.45m of GMSL rise in a very short time, perhaps no more than one year (Table 2).

THE IMPACT OF MWPs ON COASTAL AND SUBMARINE PROCESSES

The effect of MWPs at the coast is largely unknown. In areas beyond the ice sheets at the LGM, relative sea levels lay well below present, while in glaciated areas the record at present is confined to sediment accumulations in isolation basins (e.g. Shennan, 1999) and drowned valley systems (e.g. Clark et al., 2004). However, it is becoming evident that MWPs affected submarine processes, particularly along former ice margins where sediment accumulations are often thick and varied in structure and texture so that the effects of MWPs and rising GMSL on sediments may have led to submarine mass failures. Tappin (2010) has drawn attention to the instability of sediments along former ice sheet margins and has referred to the "preconditioning" of sediments, i.e. the propensity of sediments to fail under certain conditions, citing both seismic activity and coastal processes such as waves and, notably, sea level change.

Demonstrating the likelihood of a relationship between rapid sea level rises and submarine mass failures, Trincardi et al. (2003) correlated submarine mass failures in the Tyrrhenian Sea with the time of MWP1A. Later, Maslin et al. (2004) identified concentrations in the North Atlantic of continental slope mass failures at between c.15ka. cal. BP and c. 13ka. cal. BP and between c.11ka. cal. BP and c.8ka. cal. BP, the periods of, respectively, MWP1A and the early Holocene GMSL rise including MWP1B. Much debate has surrounded the mechanisms of failure, with seismicity and the release of gas hydrates cited. Brothers et al. (2013) drew attention to sea level modulated seismicity, maintaining that rapid sea level rises can increase seismicity and submarine mass failures, citing MWP1A and MWP1B. Smith et al. (2013) have taken this further, noting that the failure of submarine sediments overlying a faulted area on the Continental Shelf and Slope at Storegga, Norway, took place virtually at the same time as the rise in sea level in that area associated with the discharge of Lake Agassiz-Ojibway, thus establishing a connection between sea level rise, seismicity and sediment destabilisation. In the cases cited above, whilst rapid rises in GMSL may destabilise sediments on the continental slope and shelf, it may not be the duration of the rise but rather the rate, given that the discharge of Lake Agassiz-Ojibway occurred over months to years (e.g. Ellinson et al., 2006). We observe that the importance of sea floor sediment destabilisation should not be underestimated given that very large areas of sea floor can often be involved, with distances of run-out of debris flows sometimes exceeding 1000km (e.g. Talling et al., 2007). The significance of MWPs in generating tsunamis from seismic activity and sediment destabilisation on continental shelves and slopes is as yet unclear, but following the identification of the Holocene Storegga Slide tsunami should be investigated.

MELTWATER PULSES AND ANTHROPOGENICALLY INDUCED GLOBAL CLIMATE CHANGE

With current AGW could future MWPs come from the drainage of subglacial water bodies from Antarctica and Greenland? According to the most recent inventories, over 400 subglacial lakes exist across the bed of the Antarctic Ice Sheet (Wright and Siegert, 2012; Siegert et al., 2016). They comprise a variety of sizes and volumes (from the approx. 250km long Lake Vostok to bodies of water less than 1km in length). They occur in a number of different topographic settings, including both lakes contained within valleys to lakes that reside in broad, flat terrain. Of these lakes volume data are only available for 4, and this includes Lake Vostok, the largest Antarctic subglacial lake yet found. The total volume of these lakes is 7232km^3. This figure can be compared with the volume of the two Lake Agassiz drainage fluxes during the 8.2ka event of 49,000 and 113,100 km^3 (Teller et al., 2002). While volume data for other Antarctic subglacial lakes are not available we can make a rough estimate of the likely size of a proportion of the remainder. Wright and Siegert (2012) show that 57 lakes are more than 10km in length, and with length:width ratios between 2:1 and 6:1. Given this we estimate that the surface area of lakes greater than 10km in length lies between 51,300 and 159,600 km^2. The total area for all lakes is likely to be much larger as this figure does not take into account the size of the 300 smaller lakes. Modelling suggests that there are many more to be found (Livingstone et al. 2013) although this is disputed Siegert et al. (2016). Uncertainties in estimating the total volume of lakes in Antarctica from such incomplete data means that we have not attempted this, but even if the average depth of these 57 lakes was 1km the total volume of water available for discharge from these biggest lakes would only be slightly larger than the largest of the two meltwater drainage events at 8.2ka cal. BP. In summary, the size and volumes of these subglacial lakes is not known in sufficient detail to assess their possible contribution to meltwater discharge, but recent research indicates that Antarctic subglacial water bodies are smaller than originally envisaged (Siegert et al., 2016). There is also increasing evidence that, across the continent, these lakes are connected by a network of subglacial drainage channels (Wingham et al., 2006; Fricker et al., 2007). If this is the case, the continued and connected drainage further limits the potential of Antarctic subglacial lakes to cause large and rapid MWPs.

Very few subglacial lakes have been found under the GIS and these are small (Palmer et al 2013; Palmer et al 2015); generally less than 10 km^2 in area. It seems likely that their creation is inhibited by the relatively lower ice thickness compared with the AIS and the correspondingly lower temperatures at the base of the GIS. It is clear that episodes of future rapid sea level rise are much more likely to come from the drainage of ice from the present ice sheets forced by AGW than from meltwater lakes.

Future MWPs can now only be caused if large subglacial lakes or proglacial lakes developed and discharged in Greenland and Antarctica, or, more likely, in the event that topographical situations exist where ice sheet recession can be accelerated. Recent studies on the West Antarctic Ice Sheet indicate that irreversible retreat of several glaciers, notably the Pine Island Glacier, is now underway (Joughin et al., 2014). Localised MWPs could be generated where receding glaciers retreat up-valley along reverse-gradient slopes landward of rock thresholds and where the rapid break-up of buffered ice shelves occurs. Several studies emphasise the importance of the collapse of buttressing ice shelves in enabling rapid ice flow off land, the decay of ice streams and presumably the increase in meltwater flow. However evidence for meltwater pulses is uncertain. Notwithstanding several studies of the pattern and timing of deglaciation after the LGM (e.g. Ingólfsson et al., 1998; Anderson et al., 2002; Hall, 2009), and evidence for the possible synchronicity of northern and southern hemisphere post- LGM climate change (e.g. Mayewski et al., 1996; Rohling et al., 2004), little observational evidence for meltwater pulses has been obtained. Thus Licht (2004) maintains that Antarctica is unlikely to have been a substantial contributor to MWP 1A from evidence in the Ross Sea and Leventer et al. (2006) find no evidence for MWP 1A from the East Antarctic margin; although McKay et al. (2008) identify accelerated retreat of the Ross Ice Shelf immediately preceding MWP 1B. However in a recent paper, Weber et al. (2014) report the first direct evidence for an Antarctic contribution to MWP1A from iceberg rafted debris. Conflicting evidence is provided by available models. Thus while Deschamps et al. (2012) model a significant meltwater contribution to MWP 1A from Antarctica and while Golldege et al. (2014) model evidence of accelerated ice sheet recession, particularly coincident with Meltwater Pulse 1A, Mackintosh et al. (2011) believe that Antarctic Ice Sheets made an insignificant contribution to GMSL rise at the time of MWP 1A, and Liu et al. (2015) cannot either support or refute a significant Antarctic contribution to MWP1A.

The scientific consensus view by IPCC (Church et al., 2013) is that AGW and increasing GMSL rise could increase the rate of ice recession from Greenland and Antarctica, but it is highly unlikely that MWPs on the scale of those that occurred after the LGM would occur in the foreseeable future. Even if they did their impact on the long-term sea level rise it would likely be modest. However, more recent and alarming views are presented by Hansen et al. (2016) and deConto and Pollard (2016) who suggest that multi-metre rises in GMSL could occur by the end of this century driven by catastrophic breakup and drainage of considerable parts of the present ice sheets. They argue that the Atlantic Meridional Overturning Circulation and Southern Ocean Meridional Overturning Circulation is slowing and they estimate will shut down this century, increasing ocean stratification and concentrating warm water in the vicinity of ice shelves and floating ice margins. This will rapidly melt these, producing the conditions for ice sheet collapse. Resolution of the contrasting views from IPCC and those by Hansen et al (2016) and deConto and Pollard (2016) is clearly of enormous policy interest for climate change adaptation and in support of rapid climate mitigation strategies.

CONCLUSION

Much remains to be understood about the meltwater pulses that marked the GMSL record after the LGM. In this paper we have listed a sequence of 11 events. At present, some have only been identified locally or regionally whilst others are believed to be present globally. Such differences may not reflect reality since the sea level rises involved may have been masked by the overall GMSL rise at the time, and at present, available methodologies and techniques may be insufficient to identify the extent of a MWP. This may have given rise to the concept of a "stepped" rise in GMSL (e.g. Liu et al., 2004; Gornitz, 2013) since evidence for falls in GMSL after the recorded peaks of MWPs is equivocal. It may also be the reason for the different views on the extent of MWP1B, because this event was only one of several events contributing to the early Holocene sea level rise, when GMSL was rising at a noticeably more rapid rate than during the preceding Younger Dryas (see Smith et al., 2011).

The contributions to a given MWP are uncertain, although the break-up of ice shelves must be a major factor. The contribution of lake discharges is probably confined to ice-marginal lakes, rather than pluvial lakes, and is minor, amounting to less than around 1.2m of the c.125m of the rise in GMSL since the LGM. However, the effect may be significant where a rapid flux of water is involved: the rate of discharge, whether from ice sheet disintegration or proglacial lakes, may be important in terms of effects on the sea floor.

Perhaps the greatest contribution from an understanding of the volume, age, extent and impact of MWPs is in studies of GMSL. Guided by increasingly sophisticated models we are almost intuitively led to the assumption that the rise in GMSL after the LGM is smooth with few inflexions. Whilst this may be an acceptable general conclusion, it is far from reality in a given coastal area, where variations on the scale of even a few decimetres, which can be magnified in a locally variable coastline, may be of considerable significance for coastal processes and, by extension, for coastal societies and economies.

**Neil...the paragraph below is the previous Conclusion. Compare with paragraphs above in red. Which is better do you think?

We show that drainage of the known post-glacial lakes in total produced less than around 1.2m of the 125m of GMSLR since the LGM.   This figure is computed relative to present sea levels, and would have been higher relative to the lower sea levels (and therefore less extensive area of the oceans) in late glacial times. The figures of enhanced sea level rise during MWP events must also be seen against the background of a secular rise in GMSL which reduces their significance. We also caution that 'jumps' in GMSL may also represent decanting of the oceans into low-lying coastal regions as rising sea levels interact with an uneven global hypsometry rather than increases in water volume to the oceans. This point deserves closer modelling. The bulk of GMSL rise after the Last Glacial Maximum was driven by ice sheet decay, probably in conjunction with increased iceberg production and the impact of submarine topography as the sea level rose. We also speculate that MWP 1B was not an independent event but an element of the rapid secular sea level rise associated with the Early Holocene Sea Level Rise between 11200 and 11500 cal. BP. Finally, only MWP 1A0, MWP 1A and the drainage of Glacial Lake Agassiz at around 8.4 cal BP appear to have had a global signature; the other MWPs were only of local or regional importance.

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Figure Captions

Figure 1. Ice-volume GMSL equivalent during the Last Glacial Maximum (LGM) and Late Glacial times (from Lambeck et al., 2014). Shown are major climate events in this period including Heinrich events H1 to H3, the Bølling-Allerød warm period (B-A), and the Younger Dryas cold period (Y-D)] as well as the timing of MWP-1A, 1B, and the 8.2 ka BP cooling events.

Figure 2. Location of LGM and Late Glacial lakes, and places mentioned in the text.

Figure 3 Location of Glacial Lake Missoula and flood direction.

Figure 4 Location of Glacial Lake Agassiz-Ojibway.

Figure 5 Baltic Ice Lake and Ancylus Lake.

Table 1. Published Post-LGM MWPs. LIS: Laurentide Ice Sheet (Includes Innutian ice Sheet); CIS: Cordilleran Ice Sheet; AIS: Antarctic Ice Sheet; SIS: Fennoscandian Ice Sheet; MWP: Meltwater Pulse; GMSL: Global mean sea level.

Table 2. Published rapid proglacial and pluvial lake discharges in excess of $100km^3$ directly into the ocean in date order after the LGM.

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